Influence of rain re-evaporation on Pacific rainfall patterns in an AGCM DRAFT VERSION !!

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Short title:

Abstract. AGCM experiments show that parameterized rain re-evaporation has a large impact on simulated precipitation patterns. Specifying strong re-evaporation rates in the model leads to a better overall agreement of simulated precipitation with observations. Weak re-evaporation leads to the formation of a "double ITCZ" during the northern warm season, and an insufficiently vigorous south Pacific convergence zone (SPCZ) during the northern cold season. This sensitivity of simulated precipitation to re-evaporation may be related to differences in midtropospheric water vapor transport in the ITCZ/SPCZ complex of the central and western Pacific.

1. Introduction and Background

For most of the year the monthly-average precipitation over the Pacific ocean is dominated by a single band of intense rain rates centered between 8N and 10N, the so-called intertropical convergence zone (ITCZ). The dynamical reasons for this asymmetry in nature are not known. In atmospheric global circulation models (AGCMs) good simulations of the mean precipitation over the tropical Pacific are difficult to obtain. A frequent bias in AGCMs is the formation of a spurious second ITCZ in the southern hemisphere (e.g.; Meehl and Arblaster, 1998). While nature does show hints of a southern ITCZ over the Pacific, particularly during March through May (Zhang, 2001), this feature in AGCMs is usually too strong and persistent, lasting through the northern warm season June-September. The occurrence of double ITCZs in AGCMs leads to large rms errors in simulated precipitation, since it represents spurious rearrangement of the most intense precipitation on earth. Connections between double ITCZs and other AGCM simulation biases have not been conclusively established. However, it is clearly of concern to climate modelers, if AGCMs are producing large errors in the horizontal distribution of atmospheric latent heating. Finally, the wide distribution and similar structure of this bias in a variety of AGCMs suggests a the existence of a shared misunderstanding in current implementations of convection parameterizations.

Other researchers have examined mechanisms that control the location of ITCZs in the tropics. Chao (2000) and Chao and Chen (2001) propose a competition bewteen rotational deflection of convergent winds and wind driven surface evaporation @?? to explain the location of ITCZs close to, but not on, the Equator. As f increases away from the Equator, Coriolis effects can induce larger surface wind perturbations which in turn lead to increased surface evaporation, but at the same time, low-level convergence induced by heating becomes weaker. The role of high frequency transient motions in determining the location of the ITCZ has been studied extensively (e.g.;

Holton et al., 1971; Lindzen, 1974; Hess et al., 1993). Hess et al. showed that the character of both transients and precipitation in a zonally and equatorially symmetric aquaplanet depended strongly on the convective parameterization employed. With moist convective adjustment a single equatorial ITCZ formed, coupled with weak, poorly organized easterly wave activity. With a CAPE-based, mass flux scheme double off-equatorial ITCZs formed near 10°S and 10°N, along with intensified easterly wave activity. Philander et al. (1996) and Li (1997) examine the origin of the observed north-south asymmetry in ITCZ location using coupled ocean/atmosphere models. They find that weak initial asymmetries in atmospheric circulation arising from factors such as continental distribution, can be amplified by air-sea interactions, which favor one ITCZ over the other. Recent AGCM simulations over a "swamp"-planet, i.e., a mixed layer driven by surface fluxes only, obtain a single, but relatively diffuse, equatorial ITCZ (M.-I. Lee, personal comm.).

This study will not focus on the fundamental processes which determine ITCZ location in idealized settings. The simulations conducted here are over observed, time-varying, SST distributions (Reynolds, 1988). The AGCM used in this study (NSIPP-2.0) employs a CAPE-based, mass-flux convection scheme - Relaxed Arakawa-Schubert (RAS) (Moorthi and Suarez 1992). We expect that this model would produce off-equatorial, double ITCZs given equatorially symmetric, fixed SST forcing as in Hess at al. (1993). Nevertheless, when forced by observed SSTs, NSIPP-2.0 exhibits an interesting sensitivity in ITCZ structure to the strength of parameterized rain re-evaporation. As re-evaporation increases, the tendency for the model to form a spurious southern ITCZ decreases. The previous version of the NSIPP AGCM (NSIPP-1) (Bacmeister and Suarez 2002) exhibited similar sensitivity. However, separate treatments of re-evaporation for large-scale and convective rain in NSIPP-1 complicated the interpretation of the results (unpublished). Although this sensitivity has been useful in empirical "tuning" of the NSIPP AGCM to optimize precipitation simulations, the physical origin of the sensitivity was not understood. Unfortunately, anecdotal evidence from other modeling groups suggests that this straightforward sensitivity is not universal (I. M. Held, personal comm.), and also that other sensitivities may exist to parameters such as cumulus friction. In this study we use NSIPP-2.0, which employs a single parameterization of re-evaporation, in an attempt to isolate to the processes connecting rain re-evaporation to the seasonal mean rainfall patterns in the tropical Pacific. We hope that this will shed light on the general problem of tropical precipitation modeling, and suggest reasons to explain the variety of parameter sensitivities exhibited in AGCM rainfall simulations.

The paper is organized as follows. Section 2 provides a description of the AGCM used in this study. The parameterization of rain re-evaporation is described in some detail. Section 3 outlines the AGCM experiments performed. Section 4 presents the basic sensitivity of the model simulations to re-evaporation. Seasonal mean fields are shown, as well as some analysis of high frequency transients, thermodynamic vertical profiles, and simulated atmospheric re-evaporation rates. Section 5 describes an analysis of the water vapor budget in domains around the southern and northern Pacific ITCZs. This analysis reveals interesting differences in free-tropospheric water vapor transport that depend on rain re-evaporation. Section 6 discusses results from an experiment with added drag in the tropical free troposphere. This experiment is intended to isolate the role of transport in the maintenance of the southern ITCZ. The behavior of high frequency transients in our simulations is discussed in Section 7.

2. Model Description

We use a preliminary version of the NSIPP-2 AGCM (NSIPP-2.0) for this study. NSIPP-2.0 was developed from the NSIPP-1 AGCM, which was documented in

Bacmeister and Suarez (2000) and Bacmeister and Suarez (2002). Simulated seasonal means and responses to interannual SST variation in NSIPP-1 were both in good agreement with meteorological analyses (e.g.; Schubert et al., 2001, 2002). The significant modifications to NSIPP-2.0 and NSIPP-1 involve the cloud, boundary layer, and convection schemes. These include introduction of a prognostic cloud scheme in place of the Slingo (1987)-type diagnostic scheme used in NSIPP-1, as well as a simple moist boundary layer entrainment scheme, which is called in addition to the existing first-order dry turbulence parameterization of Louis et al (1982). These modifications were aimed at improving the models simulation of subtropical marine stratus decks. Since they have little impact on the ITCZ sensitivities examined in this study, they will not be described further here.

The dynamical core of NSIPP-2.0 is the same as in NSIPP-1 and is described in Suarez and Takacs (1996). Radiative effects in NSIPP-2.0 are parameterized using the approach of Chou and Suarez (1992). Land surface effects a parameterized according to Koster and Suarez (1996), and orographic wave drag is treated according to Zhou et al. (1996).

2.1 Rain re-evaporation

A significant change in NSIPP-2.0 is a more consistent treatment of large-scale and convective rain re-evaporation. In previous versions of the model, convective re-evaporation was based on microphysical expressions from Sud and Molod (1988), while large-scale rain was simply re-evaporated to grid box saturation at each time step. In NSIPP-2.0 the same formulation is used for both large-scale and convective precipitation. Microphysical expressions for rain re-evaporation are typically very complex since integrals over the Marshall-Palmer distribution (Marshall and Palmer, 1948) are performed at each step of a calculation involving droplet radii and fall speeds (e.g; Lin et al., 1983). These expressions are clearly preferable in high resolution calculations with prognostic precipitation species. However, in an AGCM it can be difficult to reconcile such expressions with the significant uncertainties in subgrid scale variability as well as with the large time steps ~1000 s typical of AGCM physics parameterizations.

We have adopted a compromise "single mode" approach, in which a representative droplet size r_p and fall speed is determined from a MP-distribution for a number of "showers" within each grid box. The preliminary version of the model used here uses only two such showers - large-scale (LS), and convective (CN). First, 3D precipitation fluxes $\mathcal{P}$ are estimated from the mixing ratios of precipitating condensate q_p according to:

$$\mathcal{P}^*_{(LS,CN)l} = \beta^{-1}_{(LS,CN)} \left(\frac{\rho_l \; \Delta z_l \; q_{p(LS,CN)} \; l}{\Delta t} \right) + \mathcal{P}_{(LS,CN)l-1} \tag{1}$$

where l represents a vertical level and $l-1$ is the level above. This assumes precipitating condensate is cleared from each grid box in a single time step. The parameter β represents the fractional area of each gridbox covered by the showers. We use β=1 for both LS and CN rain, i.e. we assume that precipitation is distributed uniformly through the gridbox. The second assumption is clearly suspect in the case of convective precipitation, but serves a useful baseline for future refinement.

Once $\mathcal{P}_{(LS,CN)}$ have been estimated, we use the third moment of the MP distribution to determine typical droplet sizes and fall speeds for the LS and CN precipitation streams. These are then used to calculate microphysically-based rain evaporation amounts,

$$\delta q_p = \alpha_r Ve(r_p) \frac{1-U}{\rho_w (A+B) r_p^2} \left(\frac{\Delta z}{w_f(r_p)} \right) \; q_p. \tag{2}$$

Here α_r is an empirical nondimensional parameter that modulates the strength of rain re-evaporation. U is the environmental relative humidity, assumed here to be the gridbox value. The quantity $Ve(r_p)$ is a ventilation factor (e.g.; Liou, 1992), typically between 1 and 5, that accounts for enhancements in evaporation due the air flow past the falling drop. A and B are temperature, pressure and humidity dependent microphysical quantities (e.g.; Liou, 1992; Del Genio et al., 1996) The quantity $\frac{\Delta z}{w_f(r_p)}$ is simply the amount of time spent by a drop in the layer. We assume this is shorter than the typical physics time step in our model 1800 s. For very weak showers and thick mid-tropospheric layers it is possible that this will not be true. However, no provision is currently made for this possibility other than an overall restriction on δq_p to values $\leq q_p$.

After δq_p is determined from (2) $\mathcal{P}_l$ is updated by using MAX$[q_p - \delta q_p, 0]$ in place of q_p in (1) and the calculation proceeds to the next level down $l+1$. This approach is crude, but the connections to AGCM approximations are clear, and most essential microphysics are included.

2.1 Autoconversion

The model used here still employs two distinct calculations for autoconversion. For convective rain, autoconversion is represented as a contribution to the vertical gradient of condensate in the steady-state plumes (cloud-types) invoked by RAS

$$\partial_z q_{c_{RAS}} = \ldots + (w_u \tau_{\mathcal{A}})^{-1} q_{c_{RAS}}$$

where $q_{c_{RAS}}$ is the condensate profile within an individual RAS plume, w_u is an updraft speed estimated by integrating the buoyancy force in the vertical, and $\tau_{\mathcal{A}}$ is constant time scale taken to be 1000 s. For large scale autoconversion a temperature and q_c dependent similar to that in Sud and Walker (1999) is used.

In practice, the choice of the convective autoconversion constant $\tau_{\mathcal{A}}$ has little impact on the overall precipitation simulation. It has a large impact on the partition between large scale and convective precipitation. With the value used in the experiments here, a global mean of convective to total precipitation $\frac{\mathcal{P}_{CN}}{\mathcal{P}_0}$ between 0.25 and 0.35 results. However, in the tropics most of the "large-scale" rain originates in detraining anvils. So, as long as no distinctions in the value of β in (1) are made, the distinction between large scale and convective rain is largely semantic.

3. Experimental Setup and Analysis Procedure

In this study we will discuss results from 3 baseline experiments which are identical except for the value of α_r used (Table 1) . These experiments were initialized on June 1 1989 using initial conditions from an AMIP style run using the previous version of the NSIPP AGCM, and allowed to run through December 31 1991. Horizontal resolution of 2.0×2.5 with 40 unevenly spaced σ–levels in the vertical were used for all experiments. Extensive suites of diagnostic outputs were saved from each experiment, including full 3D, daily-averaged wind, vertical motion, moisture and temperature on σ–surfaces. Most of the individual moisture tendency terms were also stored daily, including the tendency due to rain re-evaporation $\mathcal{R}$

3.1 Fictitious drag experiments

In addition to the 3 baseline experiments, we also performed a number of idealized experiments to illustrate the role played by water vapor transport in ITCZ dynamics. These were motivated by the results of water vapor budget analyses of the baseline experiments (Section 5). We will discuss the results of one of these -experiment DM, in which a fictitious drag $\mathcal{D}_m$ was introduced into in the meridional momentum equation:

$$\mathcal{D}_m = -\frac{G(\phi,\sigma)}{\tau_m} v \tag{3}$$

where the drag time scale τ_m is chosen to be 6 hours. G is a function of latitude ϕ and σ given by

$$G(\phi,\sigma) = \begin{cases} \exp\left[-\left(\frac{\phi+8^\circ}{10^\circ}\right)^2\right] & \sigma \leq 0.8 \\ 0 & \sigma > 0.8 \end{cases} \tag{4}$$

This drag thus acts primarily above 800 mb in the southern tropics. The experiments performed and their shorthand designations are summarized in Table 1.

4. Basic Model Sensitivity to Re-evaporation

4.1 Mean seasonal climate

July-August (JJA) 1990-91 seasonal mean precipitation fields for experiments B1, B2, and B3 are shown in Figure 1, along with observational estimates of precipitation rates from CMAP (Xie and Arkin, 1997). The results illustrate the important climatological control exerted by the re-evaporation strength in the NSIPP AGCM. Exp B1 (Fig. 1a) tends toward a "double ITCZ" configuration, with precipitation rates in excess of 8 mm d^{-1} extending in a narrow, zonally-aligned band along 10S well into the central Pacific. This bias with respect to observations is most pronounced during the northern warm season, roughly April-November. During December-February (DJF, not shown) all 3 baseline experiments do a reasonable job of simulating precipitation in the tropical Pacific. A NW-SE tilting south Pacific convergence zone (SPCZ) is present south of the Equator between 150°E and 150°W, although in B1 it is less well developed, and more zonally-aligned, than in B2 and B3. During JJA, in the warm pool region (120°E-150°E, Eq-15°N) and also in the central, tropical Pacific (150°E-120°W), precipitation in the experiment with strongest re-evaporation (Exp B3, Fig. 1c) appears closest to the Xie-Arkin climatology. In B1 and B2 the southern Pacific ITCZ appears to grow at the expense of the South Pacific Convergence Zone (SPCZ) and near equatorial, warm pool precipitation.

Total precipitable water (Fig. 2) also changes as re-evaporation strength varies, with the largest global mean value associated with the strongest re-evaporation. In general the experiments with significant re-evaporation B2 (Fig. 2b) and B3 (Fig. 2c) again appear to be in closer agreement with the observational estimate, in this case SSM/I total precipitable water (Alishouse et al., 1990) shown in Figure 2d. Experiment B1 is substantially drier than the observational estimate. Peak values over the Pacific warm pool in B1 are ~ 45 kg m^{-2}, where the observations show values over 55 kg m^{-2}.

Many other aspects of the model's mean seasonal climate also depend on the choice of α_r, including top-of-atmosphere (TOA) radiative fluxes, ocean surface wind stresses, and mean cloudiness (not shown). Some of these changes appear to be related to changes in the model's atmospheric water vapor distribution, and some e.g. wind stress changes, may be related to the large differences in the horizontal distribution of convective heating that occur as α_r varies.

4.2 Precipitation Variance (High-Frequency transients)

High frequency easterly disturbances have been proposed as the origin of off-equatorial rainfall maxima in the atmosphere (e.g. Hess et al. 1993). We will not attempt a detailed analysis of this mechanism in our simulations. However, it is worth establishing the degree to which high frequency transient motions vary with α_r in our model. Figure 3 shows the mean variance for May-November 1990-1991 in precipitation and vertically integrated boundary layer divergence ($\approx\omega_{850}$). Model fields were high-pass filtered with a Lanczos filter (Duchon, 1979) to eliminate modes with time scales longer than 31 d. From Figure 3 it is evident that there have been quantitative changes in the high frequency variability. Both precipitation and ω_{850} variance are higher in Exp B1, in the region of the southern ITCZ (10°S,150°E-120°W). In the SPCZ region (northeast of Australia) and in the northern warm pool region (10°N-20°N, 120°W) variance is higher in Exp B3. This simply reflects the mean distribution of precipitation in each experiment.

Other subtle differences exist between the space-time spectra of high frequency motions in Exp B1 and B3. However, we do not believe these differences are of fundamental importance in producing the changes in precitpitation between Exp B1 and Exp B3. Nevertheless, this issue will be revisited in Section 7.

4.3 Atmospheric water vapor diagnostic fields

During all of the experiments discussed here, a suite of atmospheric water cycle diagnostic fields was saved during model execution. These diagnostic fields included

3D daily averages of the water substance conversion terms in our single condensate phase, prognostic cloud scheme. These are conversion of vapor to cloud condensate, conversion of cloud condensate to precipitation, and finally conversion of precipitation to water vapor, i.e., re-evaporation $\mathcal{R}$. Figure 4 shows the 1990-1991 JJA-average, vertical mass-weighted integral of the rain re-evaporation tendency $<\mathcal{R}>$ in mm d^{-1} from experiment B3 (Fig. 4a), and the proportion of $<\mathcal{R}>$ to the total precipitation generated in the column, i.e., ϵ_1=$<\mathcal{R}>/(<\mathcal{R}>+\mathcal{P}_0)$ where $\mathcal{P}_0$ is the precipitation rate (Fig. 4b). The figure illustrates the high variability of $<\mathcal{R}>$ and ϵ_1, as well as the surprisingly high values of these quantities. Values for ϵ_1 are typically over 0.5 in the tropics. The global mean value of $<\mathcal{R}>$ for this period is 2.7 mm d^{-1}, which is nearly equal to the globally averaged precipitation rate for the same seasonal average 3.1 mm d^{-1}. By contrast for the same period in B1 (not shown), ϵ_1 is everywhere less than 0.2, and the global mean value of $\mathcal{R}$ is 0.19 mm d^{-1}.

An interesting feature of $<\mathcal{R}>$ in Fig. 4a is its resemblance to rain rate in a double ITCZ regime. It is tempting to conclude from this that the double ITCZ is eliminated by simply evaporating falling rain. This simple view is only partly correct, in the sense that, re-evaporation needs to occur in conjunction with removal of the re-evaporated water vapor from the region of the southern ITCZ. In Section 6 it will be shown that even in experiments with strong re-evaporation, in which dynamical transport of water vapor is suppressed, a southern ITCZ comparable to that in Figs 1a can form.

4.4 Profiles

In order to facilitate comparison of vertical structures, we calculate average profiles in a region bounded by 150W and 170E on the east and west, and by 14S and the Equator in the north-south direction (Box S). As shown in Figure 5, this volume contains most of the spurious southern ITCZ when it forms in experiment B1. We also examine a corresponding volume bounded by the same meridians on the east and west, but by the Equator and 14N in the north-south direction (Box N). These regions will also be used in the next section as control volumes for water vapor budget analyses.

Figure 6 shows profiles of specific humidity in Box S for Exps B1, B2, and B3. Large differences are obvious above the boundary layer, with B1 2 to 3 g kg^{-1} drier than B3 at around 800 mb. The profile from B1 is in the best agreement with NCEP re-analysis profiles (Kalnay et al., 1995) above 700 hPa, but is drier than the analysis below. The profile from B3 is wetter than the re-analysis profile at all levels. The wet bias is most pronounced above 800 hPa. It is distressing that the experiment with the better simulation of precipitation (Exp B3) appears to have larger differences from the NCEP re-analysis. However, it is also possible that these differences are partly model driven. NCEP re-analysis rainfall for example exhibits a stronger double ITCZ than is observed. In situ profiles from the central Pacific (e.g. Yin and Albrecht, 2000) exhibit a good deal of variability depending on location and underlying SST. The mean ITCZ profile from Yin and Abrecht is plotted in Figure 6b against the q-profiles from Box N. This profile is for May-June 1979, and for a longitude range of 90°W TO 120°W. Nevertheless, the Yin-Albrecht profile suggests a dry bias in NCEP re-analysis at midtropospheric levels. The fact that total precipitable water in Exps B2 and B3 is closer to SSMI observations also argues for a dry bias in the re-analysis.

The shape of the water vapor profiles in Exps B2 and B3 is more difficult to explain. The profiles show a pronounced "q-reversal" (Kloesel and Albrecht 1989) between 850 and 900 hPa. However, the height of this feature in observations is typically closer to 800 hPa and is not typical of deep convective regions (Yin and Albrecht, 2000). This structure is likely due to a combination of strong re-evaporation moistening above the MBL, and an excessively shallow, entraining MBL.

Figure 7 shows profiles of rain re-evaporation $\mathcal{R}$ for JJA 1990-1991 in Box S. Moistening from re-evaporation in Exp B3 (thick solid line) peaks immediately above the MBL (900 hPa) with values over 1.5 g kg^{-1} d^{-1}. The structure of $\mathcal{R}$ is at least partially controlled by RH (crosses), with the highest rates occurring in the warm dry air immediately above the boundary layer, and much weaker $\mathcal{R}$ within the moist MBL.

Note that an error in the prognostic cloud formulation allowed very large supersaturation (>300%) to form in cold temperatures (<210 K). This error does not impact the results presented here.

Profiles of moist static energy h and saturated moist static energy h^* for JJA 1990 are shown in Figure 8. The profiles of h for B1 and B3 reflect the large differences in q (Fig. 6). Mid-tropospheric h in B1 is correspondingly lower than in B3, while boundary layer values of h are similar in both experiments. The quantity h^* is of interest primarily because when $(h_{BL} - h^*)>0$, boundary layer parcels are buoyant. The vertical integral of $(h_{BL} - h^*)$ is closely related to CAPE. There are clear systematic differences in the h^* from B1 and B3. The profile from B3 shows greater stabilization at upper levels. The profiles of $(h_{BL} - h^*)$ (not shown) are of relatively small magnitude ($\sim 0.4\times 10^4$ J m^{-2} in B3) compared with the vertical variations in h itself ($>1.5\times 10^5$ J m^{-2}). In B1 $(h_{BL} - h^*)$ is uniformly smaller than in B3, implying smaller CAPE. This difference arises from the difference in boundary layer q between B1 and B3 which leads to lower h_{BL} in Exp B1.

It is of interest that wherever a direct comparison of thermodynamic profiles for Box S and N was shown, little difference was evident between the box-averaged profiles for a given experiment. This may not be remarkable in the case of Exp B1, where mean precipitation in Box N and S is similar (Table 2), but it is somewhat surprising that Box N and S profiles for Exp B3 were also similar.

5. Water vapor budget analysis

Straightforward analysis of model fields from simulations using different α_r yields a number of interesting features beyond the large difference in southern ITCZ precipitation. Subtle differences in the behavior of high frequency transients were obtained, as well as large differences in atmospheric water vapor content, and other thermodynamic quantities. Unfortunately, none of these differences is easily related to the change in southern ITCZ precipitation with α_r. Analysis of the water vapor budgets in Boxes N and S will at least reveal how the water vapor required to maintain high ITCZ precipitation rates is supplied. Differences in the budgets between Exps B1 and B3 may also suggest mechanisms for suppressing or enhancing precipitation in Box S which contains the spurious, southern ITCZ in Exp B1.

5.1 Box-averaged water vapor budgets

Figure 9 shows a time series of monthly-averaged, area-integrated precipitation $\overline{\mathcal{P}}$ and evaporation $\overline{\mathcal{E}}$ fluxes in these boxes from experiments B1 and B3. A striking aspect of Fig. 9 is the relative lack of structure in the evaporation time series. In both boxes and in both experiments the evaporation remains close to $4.5{\times}10^8$ kg s^{-1} for most of the 2 year duration of the experiments. Evaporation in Box N tends to be somewhat higher (0.2–$0.5{\times}10^8$ kg s^{-1}) than in Box S, and also exhibits a weak annual cycle with peak values of $\sim$$6.0{\times}10^8$ kg s^{-1} during the northern cold season. It is of particular interest that the evaporation rates in both experiments are so close. This is despite the fact that boundary layer moisture in the two experiments is different, with box average BL specific humidity in box S typically 1-2 gkg lower in experiment B1 than in B3. Thus, over fixed SSTs, evaporation rates should be higher in B1. However, surface wind speeds in the tropical Pacific are also lower in B1.

In contrast, the precipitation time-series in Fig. 9 show a good deal of variation. Seasonal variations are most pronounced for Exp B3 in Box S (thick, solid line, Fig. 9a). Values range from around $3{\times}10^8$ kg s^{-1} from April to September to over $8{\times}10^8$ kg s^{-1} during northern winter and late fall (November to March). Integrated precipitation flux for Box S, in Exp B1 (thin, solid line, Fig. 9a) is relatively constant at around 5-$6{\times}10^8$

kg s^{-1}. The difference between the precipitation time series for B1 and B3 in Box S, reflects the presence of the summertime double ITCZ bias in B1. Comparison with CMAP precipitation estimates for Box S (diamond symbols, Fig. 9a), confirms that in this region, B3 produces a much better simulation of precipitation. Large seasonal excursions in precipitation fluxes are present in the observations, and although the interannual variations in the the November-March maximum are not well captured by B3, the typical amplitude of the seasonal cycle in precipitation (5-6$\times$10^8 kg s^{-1}) is captured. Precipitation in B1 is higher than observed during April-September, consistent with the familiar "double ITCZ" bias pattern. However, during northern winter B1 exhibits a dry bias in Box S. This reflects the relative weakness of the SPCZ in Exp B1.

In Box N (Fig. 9b), the precipitation time series from both experiments are flatter on seasonal time scales. Significant month-to-month fluctuations ($\sim$1$\times$10^8 kg s^{-1}) exist in both experiments. A weak seasonal cycle is evident in the observations, with a distinct minimum in Feb-March. Overall, the precipitation in Box N is relatively constant $\sim$7$\times$10^8 kg s^{-1} throughout the year, for both experiments as well in the CMAP observations.

Time rates of change of the total water vapor mass $\partial_t\!<\!\overline{\rho q}\!>$ within boxes S and N are small compared to the fluxes in Fig. 9. Typical values for $< \rho q >$ are $\sim$10^{14} kg with month-to-month changes $\sim$10^{13} kg. Using 1 month$\approx$3$\times$10^6 s yields $\partial_t\!<\!\overline{\rho q}\!>\sim$10^7 kg s^{-1}. The sum of all fluxes into boxes S and N be similar to this amount. The barely visible bars at the bottom of Fig. 9a, show the implied $\partial_t\!<\!\overline{\rho q}\!>$ obtained from the month-to-month changes in total precipitable water within Box S. The large remaining imbalance between precipitation and evaporation, $\overline{\mathcal{P} - \mathcal{E}}$, must be compensated by vertically-integrated advective fluxes of water vapor through the sides of the boxes.

5.2 Advective water vapor transport

The horizontal transport of water vapor in σ-coordinates can be written in flux form:

$$\vec{\nabla} \cdot (\pi \vec{V}_h q)$$

where π is the surface pressure, and V_h is the horizontal wind consisting of zonal component u and a meridional component v. We will examine zonal and meridional water vapor fluxes πuq and πvq. We also examine separately the contributions of $\pi \vec{V}_h \cdot \vec{\nabla} q$ and $q \vec{\nabla} \cdot (\pi \vec{V}_h)$ to the total horizontal advective water vapor tendency in (X). These terms from the daily wind and water vapor outputs on σ-surfaces, along with daily surface pressure output, from our simulations. Some error is unavoidable in this approach since u and v are interpolated from the model's C-grid u and v-points to the grid centers, or p-points, before output. However, below it will be clear that sufficient accuracy has been obtained for the purposes of the analysis here.

The total advective water vapor flux into the boxes above is given by the sum of the vertically integrated fluxes through the four sides;

$$\begin{aligned} - \oint_{BoxN,S} \left(\int_0^1 \pi \vec{V}_h q d\sigma \right) \cdot d\vec{A} \;\; = \;\; & \int_{\text{east}}^{\text{west}} \; d\lambda \; \int_0^1 a \; cos\phi \; \pi v q \; d\sigma |_{\text{south}} \\ & - \int_{\text{east}}^{\text{west}} \; d\lambda \; \int_0^1 a \; cos\phi \; \pi v q \; d\sigma |_{\text{north}} \\ & + \int_{\text{south}}^{\text{north}} \; d\phi \; \int_0^1 a \; \pi u q \; d\sigma |_{\text{west}} \\ & - \int_{\text{south}}^{\text{north}} \; d\phi \; \int_0^1 a \; \pi u q \; d\sigma |_{\text{east}} \end{aligned} \tag{5}$$

where a is the radius of the earth. The four terms on the r.h.s will be denoted by $<\overline{\pi v q}^{\lambda}>^{N,S}_{\text{south}}$, $<\overline{\pi v q}^{\lambda}>^{N,S}_{\text{north}}$, $<\overline{\pi u q}^{\phi}>^{N,S}_{\text{west}}$, and $<\overline{\pi u q}^{\phi}>^{N,S}_{\text{east}}$ for box N or S. The total flux into the box, i.e., the l.h.s of the equation above, will be denoted by $\mathcal{A}_q^{N,S}$. Unless otherwise indicated, calculations of water vapor flux quantities are performed with daily-averaged, model outputs on σ-surfaces. Monthly or seasonal averages of these quantities, therefore include contributions from sub-monthly, transient disturbances.

Monthly-averaged, advective water vapor fluxes for Box S are shown in Figure 10. Positive values for the fluxes in Figure 10 indicate flow of water vapor into the box. First, note the close agreement between the total advective flux $\mathcal{A}_q^S$ and $<\overline{\mathcal{P}-\mathcal{E}}>^S$ confirming that errors in the advective flux calculation are small compared to the quantities of interest. In both experiments, $<\overline{\pi u q}^{\phi}>^S_{\text{east}}$ (thin dashed-line) and $<\overline{\pi v q}^{\lambda}>^S_{\text{north}}$ (crosses/thin dashed line) remain relatively constant throughout the simulation in both experiments. Water vapor enters Box S through its eastern edge at 150W, consistent with general easterly flow at low levels in the tropical central Pacific. A much weaker outflow of water vapor occurs through the northern edge. The fluxes through the western and southern edges of Box S, $<\overline{\pi u q}^{\phi}>^S_{\text{west}}$ (thin solid line) and $<\overline{\pi v q}^{\lambda}>^S_{\text{south}}$ (crosses/thin solid line), behave in a more interesting fashion. They are anti-correlated in time, with anomalous inflow of moisture form the west during November-March, coupled with anomalous outflow toward the south. In the northern warm season (April-October) the situation is reversed.

However, $<\overline{\pi u q}^{\phi}>^S_{\text{west}}$ and $<\overline{\pi v q}^{\lambda}>^S_{\text{south}}$ also exhibit interesting differences between experiments B1 and B3. The seasonal cycles of $<\overline{\pi u q}^{\phi}>^S_{\text{west}}$ and $<\overline{\pi v q}^{\lambda}>^S_{\text{south}}$ are substantially stronger in B3 (Fig. 10b). Moisture influx from the west during November-March appears to primarily responsible for supplying the anomalously strong precipitation during this period in Exp B3. Westerly moisture influx in B1 (Fig. 10a) is anemic by comparison, with $<\overline{\pi u q}^{\phi}>^S_{\text{west}}$ remaining negative for most of simulation. During April-October water vapor exits Box S through the western edge in both experiments at similar rates 2–4$\times10^8$ kg s^{-1}. During this season, which is when double ITCZ bias is strongest, the principal difference between the experiments is in $<\overline{\pi v q}^{\lambda}>^S_{\text{south}}$. In Exp B1, there is little net transport of water vapor through the southern edge of Box S. By contrast, in Exp B3 during April-October, there is removal of water vapor ($\sim$2–4$\times10^8$ kg s^{-1}) through the southern edge. Thus, when the spurious double ITCZ is present in Exp B1, the largest difference in the water vapor budget of Box S between Exp B1 and B3, is the fact that water vapor is removed through the southern edge of the box in B3, while little net flux occurs there in B1. This results in year-round advective inflow into Box S in Exp B1. In Exp B3, inflow occurs only during the northern cold season, while net advective outflow occurs in April-October.

The advective fluxes for Box N are shown in Figure 11. Here strong advective inflow ($\sim$2–4$\times 10^8$ kg s^{-1}) of water vapor is present most of the year, in both experiments. The zonal fluxes are generally stronger in Exp B3 (Fig. 11b), reflecting stronger zonal trade winds. Also, $<\overline{\pi v q}^{\lambda}>^{S}_{\rm north}$ exhibits more seasonal variablity in B3, with a distinct period of northerly inflow (June-November) and period of northward outflow (December-April).

5.3 Vertical layering of water vapor transport

It is of interest to know how the differences in water vapor transport between B1 and B3 are distributed in the vertical, and whether are primarily driven by divergent winds ($q\vec{\nabla}\cdot(\pi\vec{V_h})$) or by advection of water vapor horizontal gradients ($\pi\vec{V_h}\cdot\vec{\nabla}q$). Figure 12 shows profiles of horizontal water vapor fluxes into Box S during JJA 1990. The difference $\overline{\pi u q}^{\phi S}_{\rm west} - \overline{\pi u q}^{\phi S}_{\rm east}$ (thin solid line) is the net zonal flux of water vapor into the box, and the difference $\overline{\pi v q}^{\lambda S}_{\rm south} - \overline{\pi v q}^{\lambda S}_{\rm north}$ (thin dotted line) is the net meridional flux into the box. The sum of these is the total horizontal advective flux (thick dashed line) into the box. Below 900 hPa the total flux is near $+1.2\times 10^8$ kg m^{-2} s^{-1} in both experiments, so that the net transport of water vapor into Box S at low levels appears relatively unaffected by the changes in re-evaporation. The partition of this transport into meridional and zonal components is different, with more meridional transport in Exp B1, consistent with an expected strengthening of the meridional trade winds as precipitation becomes more zonally aligned.

However, above 900 hPa large differences in net water vapor flux are apparent. The net flux in Exp B3 becomes negative implying removal of water vapor from Box S in the free troposphere. This removal of water vapor is accomplished by net meridional transport, which is strongly negative $\sim -1\times 10^8$ kg m^{-2} s^{-1} in B3. Net meridional transport is also negative in Exp B1, but is much weaker. Further examination of the meridional flux profiles in B3 shows that the free-tropospheric, net meridional flux is dominated by export of water vapor across the southern edge of Box S.

Another interesting feature of the water vapor transport profiles from Exp B3, is the partition between the divergent and gradient components ($q\vec{\nabla}\cdot(\pi\vec{V_h})$ and $\pi\vec{V_h}\cdot\vec{\nabla}q$ resp.). Although the shape of the net flux profile is clearly determined by $q\vec{\nabla}\cdot(\pi\vec{V_h})$ (solid circles) there is a nearly constant and substantial negative contribution from $\pi\vec{V_h}\cdot\vec{\nabla}q$, as indicated by the the difference the total flux profile and the $q\vec{\nabla}\cdot(\pi\vec{V_h})$ profile. In Exp B1 this difference is generally smaller (above 900 hPa) and not systematically negative. Normally the term $q\vec{\nabla}\cdot(\pi\vec{V_h})$ is assumed to overwhelmingly dominate net water vapor transport in the tropics. It appears this assumption is marginal at best in Exp B3.

6. Fictitious drag experiment

The analysis of the simulated water vapor budgets from Exps B1 and B3 suggests that mid-tropospheric water vapor transport plays a key role in suppressing the formation of a spurious, southern ITCZ during the northern warm season. In Exp B3 Meridional transport above 900 hPa removes water vapor from a 1400×4000 km^2 domain (Box S, Fig. 5) including the spurious ITCZ, principally towards the south, and balances the excess of evaporation over precipitation. In Exp B1, for the same domain, precipitation exceeds evaporation with a net dynamical influx of water vapor closing the budget. A similar balance holds for both experiments in an equal-sized domain around the northern ITCZ (Box N, Fig. 5). A naive conclusion that could be drawn from the budget analyses is that reducing meridional transport out of Box S in Exp B3, would lead to an increase in precipitation. This possibility is examined in Exp DM (Table 1), which uses strong re-evaporation as in B3, but incorporates a strong fictitious drag in the meridional direction. The drag acts above 800 hPa, in a band centered on 8°S.

The altitiude restriction in the effect of the drag is meant to minimize its impact on low-level, meridional water vapor.

6.1 Basic climate

Mean JJA precipitation for DM is shown in Figure 13a. The precipitation pattern in the tropical Pacific resembles that from B1 more than that from B3. Rates of 8 mm d^{-1} extend well beyond the dateline along 8S. Mean JJA rainfall rates for Boxes S and N from Exps DM, B1 and B3 are given in Table 2. However, other model physics quantities, including precipitation features outside of the tropical Pacific, remain closer to their appearance in B3. Total precipitable water from DM is shown in Figure 14a. Over the tropical Pacific the TPW distribution pattern in Exp DM more closely resembles that from B1, but even there TPW values are more like those in Exp B3. The partition of continental rainfall to oceanic rainfall in the tropics is similar in B3 and DM in JJA and DJF (not shown), suggesting that this aspect of the precipitation simulation is controlled by the re-evaporation.

Figure 15 shows meridional cross-sections of the meridional wind averaged between 170°E and 150°W for JJA. The relatively unremarkable appearance of the flow in Exp DM (Fig. 15c) is noteworthy. We expected the drag in (3,4) to induce a large distortion in the atmospheric circulation. However, the meridional wind in Fig. 17c is quite acceptable, in fact it is closer to that in NCEP re-analysis (Fig. 15d) than the meridional flow in Exp B3 (Fig. 15b). Overall, a routine analysis of model output from Exp DM would not suggest that the physics had been distorted in any way. The large increase in the southward meridional wind, between 30°S and the Equator, for Exp B3, suggests a feedback interaction with increased moist heating in the SPCZ.

6.2 Vertical Profiles

A three-way comparison of vertical profiles, horizontally-averaged in Box S, for JJA 1990, from Exps B1, B3, and DM is shown in Figure 16. Mean profiles of of humidity

q (Fig. 16a) and moist static energy h and saturated moist static energy h^* (Fig. 16b) from DM and B3 are nearly identical despite the large difference in mean rainfall for Box S between these two experiments (Table 2). As noted earlier, q qnd h profiles for B1 are quite different from those in B3 above 800 hPa, due to a much drier mid-troposphere in B1. Profiles of h^* from B1 are also distinct from the other two experiments, with less stabilization evident above 600 hPa. Overall, these profiles suggest no obvious connection between the thermodynamic structure of the atmosphere and precipitation in the southern ITCZ. Rather, the mean thermodynamic structure appears to be tied to the choice of α_r in (2).

The mean cloud-base mass flux m_{cb} for different cloud-types (Moorthi and Suarez 1992) in RAS is shown as a as a function of cloud detrainment pressure level in Figure 16c. Below 850 hPa and above 300 hPa all three experiments share similar profiles of m_{cb}. Between 850 and 300 hPa, the m_{cb}-profile for B1 is substantially weaker (~30-150kg m^{-2} d^{-1}) than those for B3 This implies lower total convective mass flux, and a much higher convective precipitation efficiency ϵ_p in Box S for B1 than for B3 or DM (Table 3). Between 600 and 300 hPa, m_{cb} for DM is stronger by upto 50kg m^{-2} d^{-1}) than that for B3. This accounts for a mean, total convective, cloud-base, mass flux $\sum_l \overline{m_{cb}}$ in DM, that is about 100 kg m^{-2} d^{-1} stronger than in B3 (Table 3). Thus, deep convective clouds are somewhat stronger in Exp DM than in B3.

Although B1 and DM share similar surface precipitation rates, profiles of precipitation fluxes $\mathcal{P}$ (Fig. 16d) are strikingly different. The precipitation profile for B1 increases at a nearly constant rate from 200 hPa all the way to the surface, while that for DM increases rapidly between 200 and 500 hPa, and then remains nearly constant down to the surface. The shape of the $\mathcal{P}$ in B3 is similar, but the generation of precipitation above 500 hPa is weaker than in DM. Profiles of rain re-evaporation $\mathcal{R}$ in B3 and DM (Fig. 16e) are nearly identical. Thus it appears that the re-appearance of the southern ITCZ in Exp DM, is due to an increase in precipitation production above 500 hPa, over that in B3. This increase in precipitation may be related to the increased strength of deep convection in DM.

Also indicated in Figure 16d (thin lines) is the convective component of rainfall for each experiment. This refers to rain produced by RAS, through autoconversion of condensate within convective updrafts. Condensate remaining in the parcel as it detrains is added to the prognostic cloud water scheme, where it autoconverts according to the Sundquist-type formulation in Sud and Walker (1999). Figures 16f and 16g show profiles of cloud condensate production by detraining convection and statistical (RH-based) condensation, as well as, autoconversion of cloud condensate. From Fig. 16f it can be inferred that almost all of the non-convective rain in Box S originates in detraining anvils in all three experiments. The statistical source of cloud water for Exp DM shows an interesting increase over that in B3. However, integrated over the column it accounts for less than 10% of the increase in precipitation obtained in DM over B3.

In practice, the proportion of convective to total rain in NSIPP-2.0 is controlled by an empirically chosen convective autoconversion rate within RAS (Section 2). The choice of this parameter primarily affects the amount of precipitation originating in anvils vs. convective updrafts, and has little impact on the net rainfall.

Box averaged profiles of horizontal water vapor flux for Box S are shown in Figure 16g. These profiles are consistent with the argument that meridional transport of water vapor plays a key role in eliminating the spurious southern ITCZ. The net transport of water vapor in Exp DM is remarkably close to that from Exp B1, with strong net inflow of water at low levels (below 900 hPa) and weak horizontal water vapor transport above. In Exp B3, by contrast, strong outflow of water vapor above 900 hPa results in a net advective removal of water vapor from Box S despite inflow at low-levels.

6.3 Convective mass flux, rainfall and rainfall efficiency

From the profiles in Figure 16 are puzzling in that no large differences in thermodynamic structure are evident between Exps B3 and DM, yet precipitation in the two simulations is quite different. Horizontal water vapor fluxes (Fig. 16h) also show large differences between B3 and DM, with much weaker, free-tropospheric, transport of water vapor in DM. However, the mechanism through which these differences in transport could alter the precipitation in Box S is not obvious. Subtle differences are evident in the structure of convective mass flux (Fig. 16c) with Exp DM exhibiting somewhat enhanced deep convection. The profiles of stratiform autoconversion (Fig. 16g) show more production of rain at upper levels in DM. We now examine the relationships between deep convection, precipitation production, and precipitation efficiency.

Figure 17a shows deep convective mass flux, defined here as the sum of m_{cb} over all clouds detraining above 500 hPa ($\sum_{p<500} m_{cb}$), plotted against precipitation flux at the ground $\mathcal{P}_0$ for Exps B3 and DM, in Box S. Daily averages from June 1, 1990 to Aug 31, 1990 of all quantities are used in the Figure. It is clear that a different relationship between deep convective mass flux and $\mathcal{P}_0$ exists in the two experiments, with significantly more surface precipitation in DM for a given mass flux. Thus, the simple increase in m_{cb} seen in Fig. 16c does not entirely account for the increased $\mathcal{P}_0$ for Exp DM. In fact, using the linear $\sum_{p<500} m_{cb}$ vs. $\mathcal{P}_0$ relationship derived from Exp B3 shows that increased $\sum_{p<500} m_{cb}$ in DM would account for only around 30% of $\mathcal{P}_0$ increase obtained in DM. The relationship between total precipitation generated $\mathcal{P}_0 + <\mathcal{R}>$ and $\sum_{p<500} m_{cb}$ (Fig. 17b) is more similar in B3 and DM. Nevertheless, Exp DM exhibits somewhat stronger rates of precipitation generation for given amounts of mass flux than does B3. Using the linear relationship between $\mathcal{P}_0 + <\mathcal{R}>$ and $\sum_{p<500} m_{cb}$ for B3 shows that the increase in $\sum_{p<500} m_{cb}$ in DM would account for around 60% of the increase in $\mathcal{P}_0 + <\mathcal{R}>$.

The relationship between total evaporated precipitation $<\mathcal{R}>$ and $\mathcal{P}_0$ is shown in Figure 17c. Generally speaking there is a flatter relationship between $<\mathcal{R}>$ and $\mathcal{P}_0$ for DM. This is probably due primarily to the nonlinear dependence of re-evaporation in (2) on r_p. As precipitation flux increases, droplet size increases, contributing to slower re-evaporation rates. Ventilation and fall-speed increase at similar rates with r_p, so that their effects tend to cancel, leaving an approximate r_p^{-2} dependence for the evaporation rate. However, Figure 17c also shows that for similar rates of surface precipitation, Exp B3 tends weakly toward higher amounts of $<\mathcal{R}>$ than DM. From (2) we see that given similar rain rates (and thus similar r_p), differences in relative humidity can still result in different re-evaporation rates. Relative humidity is somewhat higher in DM (Fig. 17d, thin dash-dotted line) in a narrow band around the southern ITCZ (10°S-6°S).

7. Analysis of transient motions

The idea that high frequency propagating waves are responsible for the observed characteristics of the ITCZ has been discussed for over 30 years (e.g.; Holton, 1971; Hess et al., 1993; and Gu and Zhang, 2001). While data and models agree that high frequency disturbances (both eastward and westward) are embedded within the ITCZ, proof of causality, i.e., that these motions actually generate the ITCZ in a time-mean sense, has been elusive. In the aquaplanet simulations of Hess et al., zonally asymmetric motion appear to be required to form off-equatorial ITCZs. However, the satellite data analysis of Gu and Zhang suggests that the role of propagating disturbances changes in different sectors of the ITCZ.

As seen in Figure 3 there were different variance patterns for precipitation and boundary layer convergence in B1 and B3. Figure 18 shows variance of these quantities in Exp DM for May-November. A distinct signature of the southern ITCZ is evident in the precipitation variance (Fig. 18a). The magnitude and shape of this variance feature is similar to that in Exp B1 (Fig. 3a). Elsewhere, the precipitation variance for Exp DM is generally between that of B1 and B3, except in the northern warm pool region (120°E-150°W, 10°N-20°N) where it is close to that in Exp B3. Interestingly, the variance of integrated boundary layer convergence ($\approx -\omega_{850}$) in Exp DM is closer overall to that in Exp B3, even in the region of southern ITCZ.

Further differences in space-time power spectra and multiple field, cross-correlations exist. We believe that, while interesting, these differences are not causally responsible for the appearance or disappearance of the spurious southern ITCZ in our simulations. Nevertheless, given the wider interest in the interaction of tropical transients with precipitation, we present a brief analysis of these features in our model.

7.1 Transport

The role of transients in the net water vapor budget for Box S appears to be minor. Figure 19 compares compares the seasonal zonal and meridional water fluxes calculated using daily model output and then averaged in time, with the corresponding fluxes calculated using seasonal means of u, v, q, and π. The difference between these is the net transport across the edges of Box S by sub-seasonal transients. As can be seen in the figure, the profiles are nearly indistinguishable, with the largest differences occurring for the net meridional fluxes in Exp B3 and DM. Here, transient motions lead to a minor $\sim$10% increase in the outflow of water from the box above the PBL, and a similar reduction of the meridional inflow within the PBL.

7.2 Space-Time Frequency Analysis

Two-dimensional FFTs of precipitation were also calculated along 8°N (Figure 20). The figure shows background spectra, i.e., the 2D (ω-k) FFT after repeated applications of a 1-2-1 filter (Wheeler and Kiladis, 1999). The precipitation fields were 31-day high-pass filtered before the FFT was calculated. Four 90 day periods of the filtered data were used; i) May 1, 1990–July 30, 1990; ii) July 31, 1990–October 29, 1990; iii) May 1, 1991–July 30, 1991; and iv) July 31, 1991–October 29, 1991. Power spectra from these four periods were averaged to obtain the results in Figure 20. The results at periods shorter than 31 days (f>0.03 cpd) look qualitatively similar to those shown by Gu and Zhang (2001) for OLR in the ITCZ. The background spectrum is red, with power concentrated in westward moving disturbances (k<0). It is difficult to say whether any of the experiments is qualitatively closer to the Gu and Zhang result. Determination of spectral peaks above the background was not attempted due to the short length of the experiments.

Some interesting differences between experiments are evident. More power in eastward moving disturbances is found in B3 and DM than in B1, suggesting that these modes depend in some way on strong re-evaporation. The speed of the dominant westward moving transients is somewhat slower in B1 than in B3, and there is more power at higher zonal wavenumbers. In this regard, Exp DM appears to lie between B1 and B3. Overall, however the differences in the power spectra of high-frequency transients can be described as subtle.

7.3 Correlation between PBL convergence and precipitation

An intriguing difference between the experiment with weak re-evaporation – Exp B1 and the two with strong re-evaporation – B3 and DM, occurs in the correlation of boundary layer convergence and precipitation. Figure 21 shows maps of the correlation between time-series of these quantities at points between 25°S and 25°N. The time-series have been high-pass filtered at 31 days. The correlations for Exp B1 (Fig. 21a) are much higher than for the other experiments. Peak values above 0.8 are found in the southern ITCZ region, with r>0.7 over broad areas of the tropical Pacific. By contrast, values in B3 and DM are generally between 0.4 and 0.5 in most of the tropics, with small areas of r>0.6 over the northern warm pool. While this contrast is suggestive of a different degree of interaction between dynamics and precipitation, its implications for tropical rainfall on climate scales is not clear. Exps B1 and DM exhibit similar seasonal mean precipitation in the tropical Pacific, yet the correlation between precipitation and BL convergence appears to be qualitatively different.

8. Discussion

Our results suggest that horizontal water vapor transport above the boundary layer is critical to suppressing the formation of a spurious second ITCZ south of the Equator during the northern warm season. When sufficiently strong outflow of water vapor exists in the free troposphere out of the domain of the southern ITCZ our AGCM cannot sustain large precipitation rates, and a second ITCZ does not form. The vigorous export of water vapor from this domain depends on having both substantial flow and high humidity in the free troposphere. In our model, large water vapor mixing ratios in the free troposphere, result when strong re-evaporation of precipitation is allowed. In two experiments that were identical except for the strength of re-evaporation, widely different precipitation simulations resulted, with a pronounced double ITCZ present in the experiment with weak re-evaporation. Although it was not discussed in this study, a similar sensitivity has existed in the NSIPP AGCM across a number of re-evaporation formulations and other moist physics changes - with weak re-evaporation always tending towards simulations with double ITCZs. Thus, we feel that this sensitivity, while it may be unique to the NSIPP AGCM, is not unique to the particular suite of moist physics parameterizations used in this study.

However, the simulations examined here also show that high specific humidity in the free troposphere is a necessary but not sufficient condition for eliminating the spurious southern ITCZ. Sufficiently strong water vapor transport out of the southern ITCZ domain also requires a vigorous meridional circulation. When this circulation is artificially interrupted, a spurious, southern ITCZ forms in our model even with strong re-evaporation. This suggests that in order for a model to exhibit the sensitivity seen in the NSIPP AGCM, stronger re-evaporation leading to the elimination of the double ITCZ, the model's tropical circulation must be strong enough in the free troposphere to effectively remove the re-evaporated water vapor from the vicinity of the spurious ITCZ.

The clearest evidence for this dynamic comes from Exp DM, which employed a fictitious meridional drag (3,4) in the southern tropics to suppress the meridional wind. While the distortion to the model's physics from the fictitious drag in Exp DM may be regarded as severe, in most respects the simulation in Exp DM is quite acceptable. Mean profiles of water vapor, static energy and other quantities in Exp DM are nearly identical to those in Exp B3, which employs the same re-evaporation strength (Table 1) However, differences in the tropical meridional between Exp B3 and Exp DM appear to be responsible for eliminating the spurious ITCZ in one of the experiments (B3) but not the other.

The meridional winds in Figure 17 suggest that our model reacts strongly to increased moist heating in the SPCZ region. This would explain the strong intensification of the southward v in the free troposphere, Eq-30°S, seen in Exp B3 Fig. 17b. We speculate that if a model does not react this way to convection in the SPCZ, perhaps due to different convective heating profiles, the circulation may be more like the one for Exp DM (Fig. 17c), and even with strong re-evaporation the spurious ITCZ will not vanish.

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